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Sediment Transport Under Ice

Robert Ettema and Steven F. Daly

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Sediment Transport Under Ice

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ABSTRACT

An essential feature of alluvial rivers and channels is that their morphology and flow-resistance behavior vary interactively with flow and sediment conditions. Depending on flow magnitude, ice covers modify the interaction, doing so over a range of scales in space and time. This report is an in-depth review of the impacts of river ice covers on sediment transport. The following topics are covered: ice-cover influences on flow distribution; sediment transport by ice (i.e., sediment included in drifting ice); sediment transport under ice; and ice influences on channel morphology. The flow distribution in channels can be substantially modified by river ice. The impacts can include raised water levels, laterally redistributed flow, reduced velocity of secondary currents, and other effects. Drifting ice can be an important transport mechanism for sediment transport, and the known pathways are described. Sediment transport under ice is described in terms of key non-dimensional parameters characterizing the dynamics of flow and sediment interaction. Finally, the extent to which the seasonal formation and breakup of ice perturbs the stability of alluvial channels in regions subject to frigid winters is described.

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PREFACE

This report was prepared by Dr Robert Ettema, Professor, Engineering Department, University of Iowa, Iowa City, and Dr. Steven F. Daly, Research Hydraulic Engineer, Remote Sensing/Geographic Information Systems and Water Resources Branch, Cold Regions Research and Engineering Laboratory, U.S. Army Engineer Research and Development Center.

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Sediment Transport Under Ice

ROBERT ETTEMA AND STEVEN F. DALY

1 INTRODUCTION

The annual cycle of ice formation and breakup influences sediment-transport dynamics in waterways. The magnitude of the influences depends on a combination of factors. Of prime importance are factors determining the amount of ice formed and the quantities of water and sediment to be conveyed.

The overall amount of ice, and ice-mass thickness, usually are governed by the cumulative period of temperature degrees below the freezing temperature of water. Water-flow rate directly determines quantities of sediment transport. Under natural conditions in many waterways, rivers, and canals, the annual cycle of ice formation is accompanied by a decline in water runoff and channel flow. Rates of sediment supply and channel transport diminish commensurately. Runoff and channel flow subsequently increase during spring thaws, and it is then that ice-cover effects on sediment transport can become significant. For flow-regulated rivers and channels downstream of reservoirs, though, ice effects on sediment transport and alluvial-channel behavior are of special interest. Substantial flows may occur while such rivers and channels are ice covered in winter.

Ice effects can occur over varying scales of time and channel length. On the time scale of months and length scale of miles of channel, for instance, ice alters the relationships among flow rate, flow depth, and sediment transport rates. As it forms, an ice cover usually increases and redistributes a channel's resistance to flow, and reduces its overall capacity to move water and sediment. In a sense, because the channel's bed roughness does not actually increase (in fact it may reduce [Smith and Ettema 1997]), the effect on channel morphology of an ice cover's presence may be likened to the effect produced by a reduction in energy gradient associated with flow along the channel. More precisely, it may be likened to a change in thalweg geometry; the additional flow energy consumed overcoming the resistance created by the cover offsets a portion of the flow's energy the channel dissipates by thalweg lengthening or bifurcation.

At the local scale, an ice cover over a short reach may redistribute flow laterally across the reach, accentuating erosion in one place and deposition in another place. Such local changes of the bed may develop during the entire cycle of ice formation, presence, and release. They may develop briefly, lasting slightly longer than the ice cover and disappear shortly after the cover breaks up. Or, they may trigger a change that persists for some time. In any event, they should be verifiable from a site investigation.

Ice may dampen or amplify erosion processes locally. Obvious dampening effects of ice are reduced water runoff from a watershed, cementing of bank material by frozen water, and ice armoring of bars and shorelines by ice-cover set-down with reduction in flow rates. Yet, ice may amplify erosion and sediment-transport rates, notably during the surge of water and ice consequent to the collapse of a large ice jam. For unregulated or wild rivers, like the Yellowstone River shown in Figure 1, as well as for regulated rivers, it has become important to understand channel response to the winter cycle of ice.

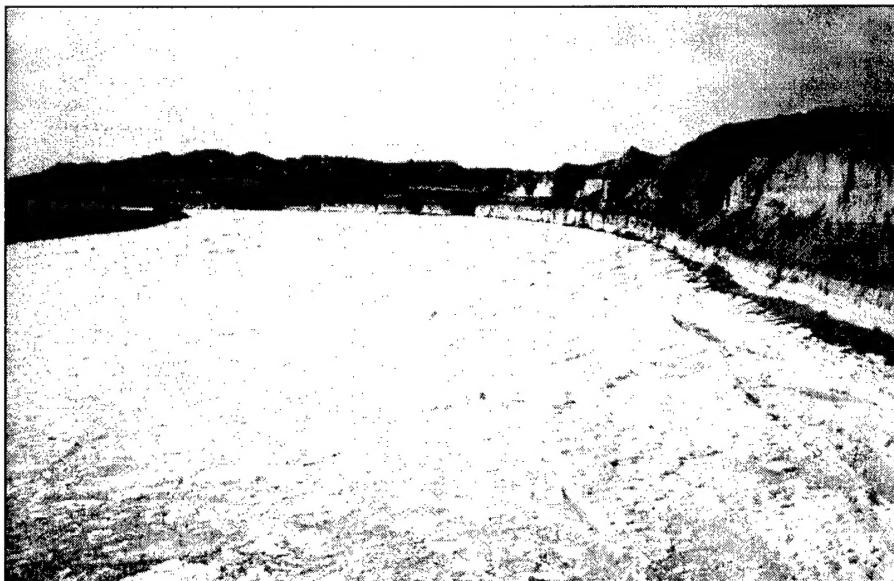


Figure 1. Yellowstone River, Montana, under an ice cover, whose formation, presence, and eventual breakup significantly influence sediment-transport dynamics, channel-thalweg location, and riverbank erosion.

This report considers the following topics related to sediment transport:

- Ice-cover influences on flow distribution.
- Sediment transport by ice (i.e., sediment included in drifting ice).

- Sediment transport under ice.
- Ice influences on channel morphology.

Sediment transport under ice is the central topic addressed herein.

The present description of sediment processes in ice-covered rivers and channels does not address the influences of permafrost on sediment transport. Permafrost is an important factor affecting riverbank and channel stability of high-latitude rivers. Lawson (1983) and Scott (1978), for example, provide some insights into of channel behavior in permafrost. Andersland and Anderson (1990) and Johnston (1981) usefully describe the geotechnical properties of permafrost.

2 ICE-COVER INFLUENCE ON FLOW DISTRIBUTION

It is useful, to briefly elaborate on the influence of an ice cover on flow distribution, which in turn influences sediment transport.

An ice cover imposes an additional resistant boundary that decreases a channel's flow capacity and vertically redistributes streamwise velocity of flow in a channel. If the cover is free-floating, it may reduce the erosive force of flow in the channel and thereby reduce rates of sediment transport. However, cover presence also may laterally redistribute flow, usually concentrating it along a thalweg. If the thalweg lies close to one side of a channel, flow concentration may locally increase bank erosion and channel shifting. On the other hand, if the thalweg is more-or-less centrally located in a channel, the cover may reduce bank erosion and channel shifting. Additionally, if the full cover were fixed to the riverbank, it may increase locally flow velocities and rates of sediment transport. The variability of flow response to ice-cover presence makes it difficult to draw simple overall conclusions about ice-cover effects on a river's bed and banks. The net effects will vary from site to site.

If the flow rate and channel slope were assumed constant, the main individual effects of a uniformly thick ice cover on a straight uniformly deep alluvial channel are as follow:

- Raised water level (ice-covered depth exceeds open water depth for the same flow rate, as illustrated in Fig. 2).
- Reduced bulk velocity of flow (discharge/flow area).
- Laterally redistributed flow (Fig. 3 and 4).
- Reduced drag on the channel bed.
- Reduced velocity of secondary currents, and altered pattern of secondary currents (i.e., currents associated with transverse circulation of flow in the channel, as shown in Fig. 5).
- Reduced rates of bed-sediment transport.
- Altered size and shape of bedforms (notably dunes).

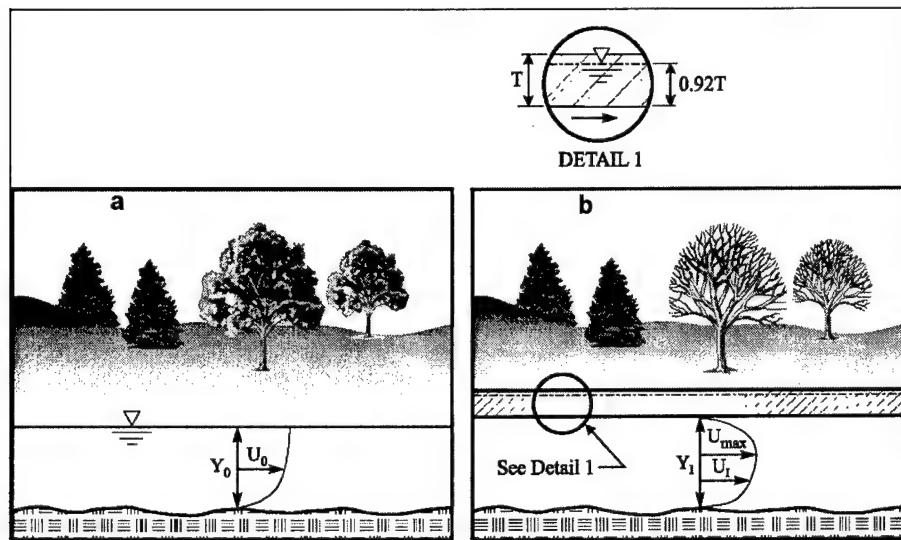


Figure 2. Ice-cover presence usually increases flow depth and redistributes flow.

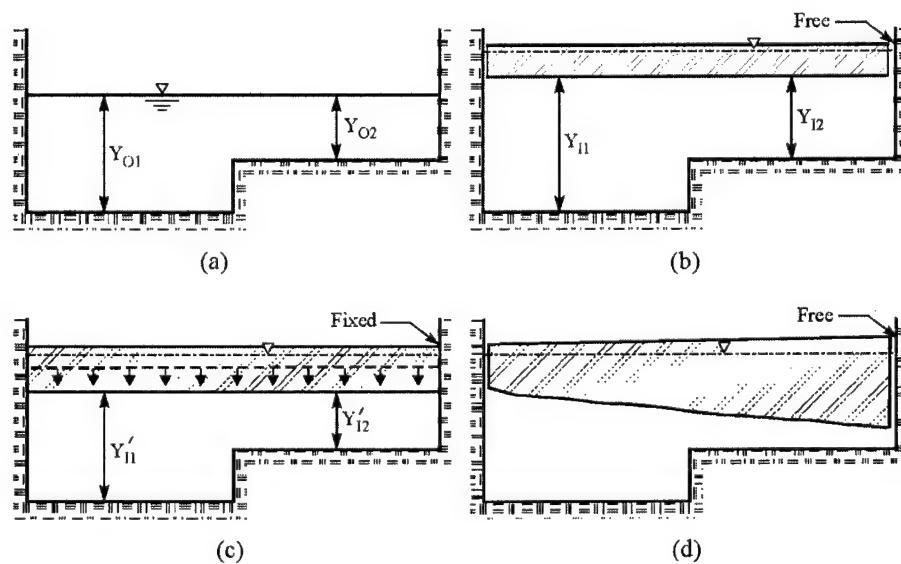


Figure 3. An ice cover may reduce open water proportions (a) of flow conveyance in lateral segments of a two-part compound channel if the cover is level and free floating (b); increase them if the cover is fixed and thickens (c); or, increase them if the cover is not uniformly thick (d).

For steady rate of water flow in an open channel (Fig. 3a), a free-floating and uniformly thick ice cover smears flow over the full channel width (Fig. 3b). However, if the ice cover were fixed to the channel banks and thickened, the reverse occurs (Fig. 3c), because flow depth reduces more in the shallower portion. Under this condition, cover presence squeezes or concentrates flow along a thalweg, where flow is deeper. If the thalweg lies close to one side of a channel (e.g., near the outer bank of a bend), such a concentration of flow may promote thalweg shifting and deepening. On the other hand, if the thalweg were located more-or-less centrally in a channel, a fixed cover may deepen or entrench the thalweg. An important further point is that the cover, by reducing flow through the shallow portion, may trigger further reductions in conveyance through the shallower portion by promoting ice accumulation (frazil slush or pans) or bed-sediment deposition, or both, there. Additional flow concentration is possible if the cover were not uniformly thick (Fig. 3d), if ice grounds on the channel bed, or if shorefast or accumulated ice develops from one or both banks.

Lateral variations in cover thickness may further concentrate flow in a channel of non-uniform depth and may override the more subtle effects to those just described for a level ice cover. Significant lateral and streamwise variations in cover thickness may occur in channels with significant variations in flow depth and velocity. Because flow velocities decrease with decreasing flow depth, velocities usually are lower in regions of shallower flows and often in the wake of flow obstructions, such as bars. Ice covers whose formation involved substantial amounts of frazil-ice slush may become thicker in regions of shallower flow. Lower values of flow conveyance in those regions also result in relatively faster bankfast-ice formation. Also, because flow velocities are lower, ice (frazil slush and ice pieces) is less readily conveyed through those regions, and is prone to accumulate. Figure 4 illustrates the accumulation of ice at a cross-section of the Tanana River, Alaska, at two times during winter (Lawson et al. 1986). That river is comparable to the lower Missouri River in flow rates, but is of steeper slope, is more braided in channel morphology, and its flow is not regulated.

Further concentration of flow is possible if an ice cover is not free to float upwards with increasing flow rate. Hydraulic analyses usually assume (e.g., Ashton 1986, Beltaos 1995) that ice covers are free floating; i.e., streamwise cracks separate the floating ice cover from adjoining bankfast ice. Actually, a cover may not always be free-floating. A stationary cover exposed to very frigid air may fuse to the channel banks. The cover then becomes constrained from freely floating up or down with changes in the flow, at least initially. Therefore, increasing flow is forced partially beneath the ice cover, initially pressurizing it and increasing flow velocities, which may locally erode the bed beneath the cover. The extent to which a flow may be pressurized beneath a cover apparently

has not yet been measured. Some evidence (Zabilansky et al. 2002) suggests that scour of the channel bed possibly may relieve the pressurized flow in alluvial channels. Very little information exists on this flow condition, especially with regard to how it may locally affect the channel bed and banks.

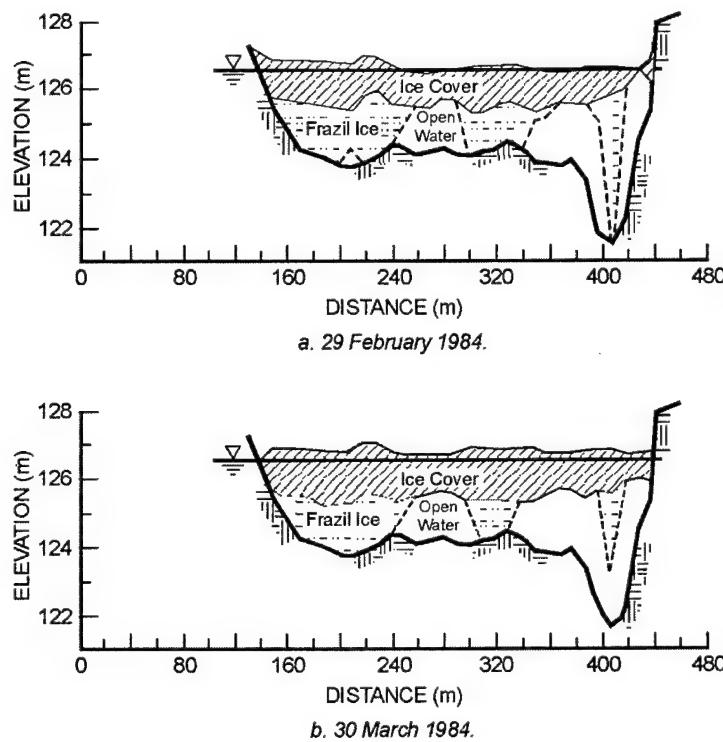


Figure 4. Non-uniform ice accumulation across a section of the Tanana River, Alaska (from Lawson et al. 1986).

For constant discharge, a free-floating level ice cover reduces bulk flow velocity and alters the vertical distribution of streamwise flow. In so doing, it usually dampens secondary currents. Cover presence reduces the centrifugal acceleration exerted on flow around a river bend, though only one study has investigated this effect (Tsai and Ettema 1994). That study found that cover presence alters patterns of lateral flow distribution in a channel bend. The two sketches in Figure 5 show the main alteration, which is a splitting of the large secondary flow spiral into two weaker spirals; owing to centrifugal acceleration acting on moving water, a large secondary flow spiral is typical of many curved channels. The presence of a level ice cover reduces radial components of velocity and lateral bed slope in channel bends, causing the bed level to rise near the outer bank. Tsai and Ettema found a reduction in lateral bed slope of about 10%. This

ice-cover effect would tend to mildly retard bank erosion in channel bends, because it may result in reduced flow velocities near the outer bank of a bend.

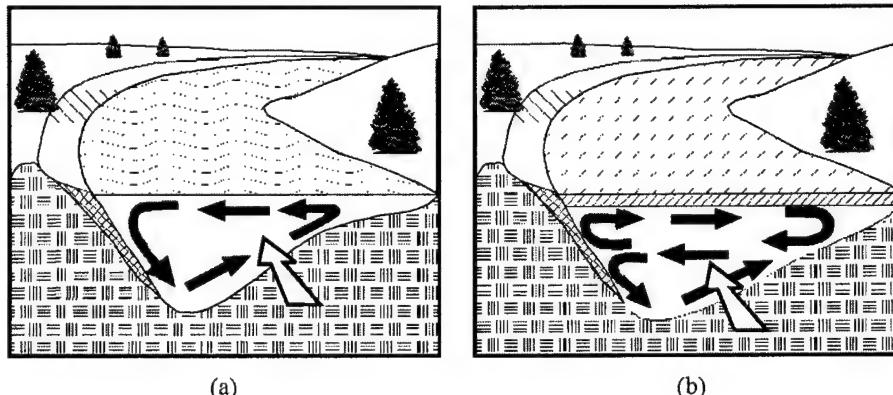


Figure 5. Ice-cover effects on spiral secondary current in a channel bend.

3 SEDIMENT TRANSPORT BY ICE

Sediment-laden ice slush and clumps of ice-bonded sediment may appear during the early stages of ice formation in certain rivers and streams subject to the winter cycle of ice formation. The ice slush and clumps usually comprise a mix of frazil ice and anchor ice that were once briefly bonded to the bed of such rivers and streams (Fig. 6). The amounts of sediment entrained and rafted with the ice slush and clumps can produce a substantial momentary surge in the overall quantity of sediment moved by some rivers and streams, though at present there are no reliable measurements or estimates of ice-rafted sediment-transport rates. Much of the entrained sediment becomes included in an ice cover, where it remains stored until the cover breaks up. Though ice rafting of sediment is known to occur (observations are reported by, for example, Barnes 1928, Wigle 1970, Michel 1971, Benson and Osterkamp 1974, Kempema et al. 1993), the engineering and environmental implications of its occurrence are largely unknown.

The short treatment given in this section limits itself to ice transport of sediment in rivers, canals, and streams. Shallow coastal (marine and lacustrine) waters in cold regions also are prone to bed-sediment entrainment and ice rafting by frazil and anchor ice. Barnes et al. (1982), Osterkamp and Gosink (1984), Reimnitz and Kempema (1987), and Kempema et al. (1986, 1993) describe coastal locations where ice entrains significant quantities of sediment. Storms in frigid weather conditions agitate coastal waters and can produce large quantities of frazil ice. The mechanisms whereby anchor ice forms in coastal waters include the same elements causing anchor ice to form in rivers and streams. The formation mechanisms for coastal anchor ice are complicated, however, by the more complex flow conditions of coastal waters and by salinity considerations in marine systems. Ice can significantly affect sediment erosion and deposition in estuaries and tidal reaches of rivers. Desplanques and Bray (1986) and Morse et al. (1999) describe the influence of ice accumulation in estuaries of the northeast portion of the Bay of Fundy. The accumulations form as ice walls from stranded ice and included sediment. The ice walls confine flow and can accentuate localized channel scour. Of particular concern in this regard is scour near hydraulic structures, such as bridge piers and abutments.

The mechanisms whereby ice entrains and transports sediment are not well understood. Also, the distances over which ice-rafted sediment typically may be transported are not really known. To date the only detailed laboratory investigations of the entrainment mechanisms are the studies reported by Kempema et al. (1986) and (1993). A handful of experiments on anchor ice formation have been

conducted, though (e.g., Tsang 1982, Kerr et al. 1998). The literature indicates that the following two mechanisms contribute to anchor ice formation.

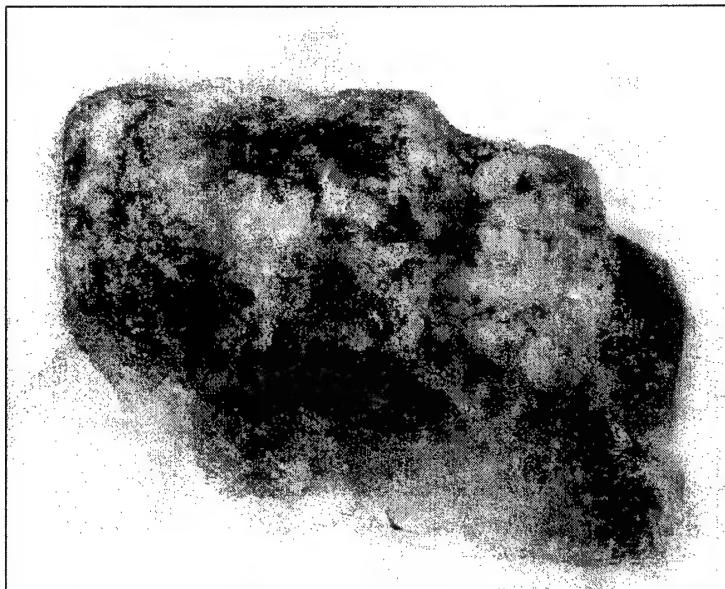


Figure 6. Sediment-laden frazil slush taken from Flat Creek in Jackson, Wyoming.

The main mechanism is frazil adhesion to bed sediment. Large-scale turbulence in comparatively shallow, swift-flowing rivers and streams can mix suspended ice crystals and flocs of active frazil across the full depth of flow. While the flow is supercooled, the frazil may adhere to bed sediment or individual boulders and accumulate as a porous and spongy mass (Wigle 1970, Arden and Wigle 1972, Tsang 1982, Beltaos 1995). Rapids and riffles are common locations for anchor ice formation (Marcotte 1984, Terada et al. 1997). Altberg (1936) reports the occurrence of anchor ice in river flows as deep as 20 m. The foregoing references report rapid rates of anchor ice growth, such that large volumes of anchor ice form in a short period.

A much less significant mechanism for sediment transport is direct ice growth on the bed or on objects protruding from the bed. Together with frazil ice, supercooled water can be mixed across the flow depth. The downdraft of supercooled water chills objects in the flow (e.g., boulders and debris of various types) and enables ice directly to nucleate and form on those objects. The resultant ice crystals are relatively small, and develop a fairly smooth and dense ice mass (e.g., Ashton 1986, Kerr et al. 1998).

The diurnal formation of frazil and anchor ice may result in repeated ice-rafting events along a river reach, each event potentially entraining substantial quantities of bed sediment. Under conditions of sufficiently frigid weather and substantial flow turbulence, extensive areas of a river's bed can become blanketed by anchor ice. As the larger sizes of sediment on a riverbed protrude more into the flow, they usually are more affected by the thermal condition of the flow than that of the bed on which they rest. Significant heat flux can occur from the sub-bed zone that is a degree of two above 0°C and supercooled flow essentially over the full flow depth of a river or stream. Consequently, larger amounts of anchor ice typically form on beds of coarser sediment. Several factors militate against extensive anchor ice formation on riverbeds of fine, noncohesive sediment. In particular, such sediments are readily lifted and thereby cannot hold a significant anchor ice accumulation (Arden and Wigle 1972, Marcotte 1984).

The laboratory studies provide interesting insights into aspects of anchor-ice formation in the presence of bedforms. They show how frazil flocs become sediment laden and lose their buoyancy as they tumble along the flume's sand bed, and eventually become included within an ice-sand clump of anchor ice. As the negatively buoyant flocs of frazil and sediment accumulate in the trough of a dune or a ripple (as illustrated in Fig. 7), they become infiltrated by sand, buried, and compressed. The resulting clumps of bonded ice and sediment may then enlarge as additional frazil flocs fuse to them, or as the clumps further grow amidst supercooled water.

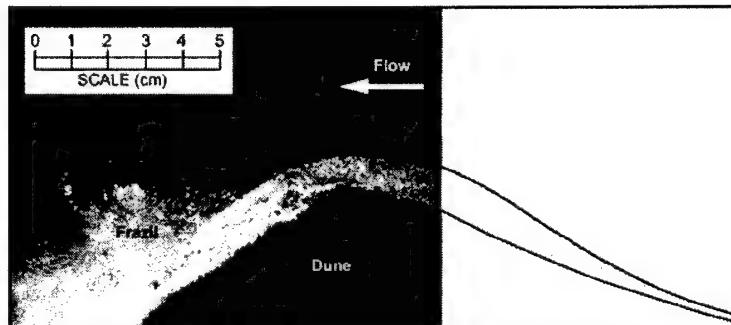


Figure 7. Frazil ice accumulated in the lee of a dune, which migrates and envelopes the frazil, eventually forming an ice-bonded clump of sand. This photo was taken from a flume experiment described by Kempema et al. (1993). (Photo taken by Ed Kempema, University of Wyoming.)

Eventually, a clump of anchor ice accumulation may attain sufficient buoyancy to lift sediment from the bed. The resulting concentrations of suspended

sediment that the ice conveys can get quite high. Kempema et al. (1986) calculate that a neutrally buoyant clump of ice-bonded sediment may contain up to 122 g of sediment per 1 L of ice and sediment. Kempema et al. (1998) measured sediment concentrations in released anchor ice masses in southern Lake Michigan of 1.2 to 102 g/L, with an average concentration of about 26 g/L. Anchor ice may move gravel and cobbles. Martin (1981) mentions an instance where anchor ice entrained and moved boulders up to 30 kg in weight. Such ice rafting can move cobbles and boulders through long reaches of deep river pools with relatively sluggish flow.

Frazil and suspended sediment may interact directly in the water column, causing suspended sediment to be included in ice slush. The details of the interactions, and the likelihood of their occurrence, are not well understood. Barnes (1928) and Altberg (1936) mention an intriguing observation that frazil-ice formation appears to remove suspended sediment from a flow; after a frazil-ice event, water seems clearer. While frazil and anchor ice form, it is possible that they may diminish bed-sediment entrainment and transport. Initially, accumulating frazil and anchor ice would bind bed sediment, thereby retarding entrainment. Also, by virtue of the ice concentrations involved, frazil ice may dampen flow turbulence, a key factor in suspended-sediment transport. Once anchor ice lifts from a bed, however, it would entrain and convey sediment, although that sediment may become frozen and temporarily stored in a floating ice cover.

4 SEDIMENT TRANSPORT UNDER ICE

Sediment-transport processes in ice-covered flow are not fully understood. Some influences of an ice cover on flow distribution are reasonably well understood, while some are barely recognized (e.g., on large-scale turbulence); few reliable methods exist for estimating transport rates or ice effects on channel stability. An essential feature of alluvial rivers and channels is that their morphology and flow-resistance behavior vary interactively with flow and sediment conditions. Depending on flow magnitude, ice covers modify the interaction, doing so over a range of scales in space and time.

Parameters

It is convenient to describe ice-cover influences in terms of key nondimensional parameters characterizing the dynamics of flow and sediment interaction. Dimensional analysis of variables associated with flow in an alluvial channel (Fig. 8) formally identifies the important parameters.

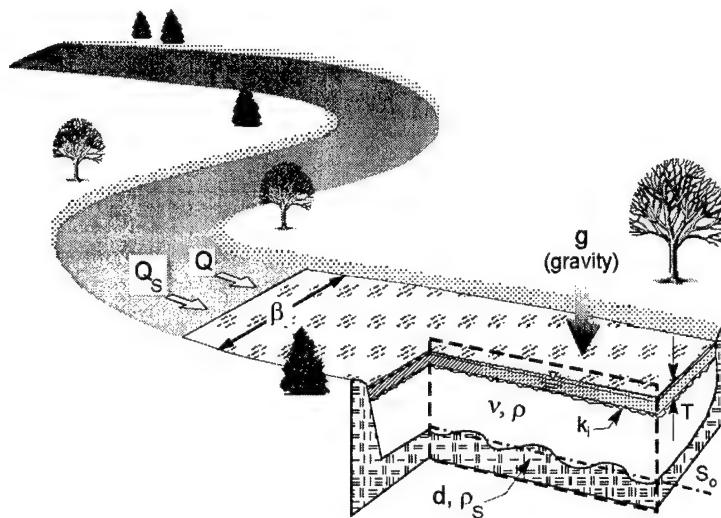


Figure 8. Independent variables usually associated with flow in an alluvial channel with an ice cover.

Typically, a dependent quantity A of a channel may have the following functional dependency for flow in a reach of comparatively wide channel comprising a bed of uniform-diameter sediment and under a uniformly thick ice cover:

$$A = f_A(Q, Q_s, v, \rho, \rho_s, d, g, S_0, B, T, k_i) \quad (1)$$

where

Q_s = Sediment discharge into reach

v = kinematic viscosity of water

ρ = water density

ρ_s = sediment density

d = median size of bed particles

g = gravity acceleration

S_0 = channel slope

B = reach width

T = ice-cover thickness

k_i = hydraulic roughness of ice-cover underside.

The dependent quantities (A) of practical concern for a channel reach are flow depth, hydraulic radius, bulk velocity of flow, flow-energy gradient, sediment-transport capacity of the flow in the reach, and possibly thalweg alignment (path of greatest depth).

Though ice-cover properties T and k_i actually also may be dependent variables, especially for ice covers formed from accumulated drifting ice, here they are treated as independent variables. Though k_i directly affects flow resistance, cover thickness, T , affects flow insofar that it is of use in characterizing cover rigidity and elevation of hydraulic grade line. The present focus is on the ways in which an existing ice cover modifies interactions between flow and bed. (See Decker and Zabilansky [2004] for a recent experimental investigation of scour under varying ice conditions.) An interesting broader discussion would be to consider how flow and bed interaction influence cover formation. That discussion might include thermal variables, such as water temperature.

In terms of non-dimensional parameters and, for convenience, considering unit discharges of water, q ($= Q/B$) and sediment q_s ($= Q_s/B$), eq 1 may restated as

$$\Pi_A = \varphi_A \left(\frac{q}{v}, \frac{q_s}{v}, d \left(\frac{g \Delta \rho}{\rho v^2} \right)^{1/3}, \frac{\rho_s}{\rho}, S_0, \frac{d}{k_i}, \frac{T}{d} \right) \quad (2)$$

in which sediment diameter, d , is used as the scaling or normalizing length. Here, sediment discharge is total sediment discharge. Also, $\Delta\rho = \rho_s - \rho$.

The second and third parameters can be combined to more usefully express sediment transport non-dimensionally as

$$\frac{q_s}{v} \left[d \left(\frac{g\Delta\rho}{\rho v^2} \right)^{1/3} \right]^{-3/2} = \frac{q_s}{\sqrt{g(\Delta\rho/\rho)d^3}} \quad (3)$$

For most situations, ρ_s/ρ is more-or-less constant (about 2.65). Whence eq 3 reduces to

$$\Pi_A = \varphi_A \left(\frac{q}{v}, \frac{q_s}{\sqrt{g(\Delta\rho/\rho)d^3}}, d \left(\frac{g\Delta\rho}{\rho v^2} \right)^{1/3}, S_0, \frac{d}{k_i}, \frac{T}{d} \right) \quad (4)$$

In eq 3 and 4, Froude number, $Fr = (q/Y)/(gY)^{0.5}$ is a dependent parameter, as flow depth, Y , is a dependent variable.

For the case of a long, rigid, and uniformly thick, free-floating ice cover, in which the significance of T/d diminishes, and eq 4 simplifies to

$$\begin{aligned} \Pi_A &= \varphi_A \left(\frac{q}{v}, \frac{q_s}{\sqrt{g(\Delta\rho/\rho)d^3}}, d \left(\frac{g\Delta\rho}{\rho v^2} \right)^{1/3}, S_0, \frac{d}{k_i} \right) \\ &= \varphi_A \left(Re, \frac{q_s}{\sqrt{g(\Delta\rho/\rho)d^3}}, D_*, S_0, \frac{d}{k_i} \right) \end{aligned} \quad (5)$$

in which

$$D_* = d \left(g\Delta\rho / [\rho v^2] \right)^{1/3}$$

and

$$Re = \frac{q}{v}.$$

Many relationships in alluvial-channel hydraulics are expressed in terms of particle Reynolds number

$$Re_* = u_{*b} d / v$$

and Shield's parameter

$$\theta = \rho u_{*b}^2 / (g \Delta \rho d)$$

here, u_{*b} = shear velocity associated with bed component of velocity distribution.

In this regard, using $D_* = (Re_*)^2 / \theta$, eq 5 more usefully is

$$\begin{aligned} \Pi_A &= \varphi_A \left(Re, \frac{q_s}{\sqrt{g(\Delta \rho / \rho) d^3}}, D_*, S_0, \frac{d}{k_i} \right) \\ &= \varphi'_A \left(Re, \frac{(Re_*)^2}{\theta}, \frac{d}{k_i}, \frac{q_s}{\sqrt{g(\Delta \rho / \rho) d^3}}, S_0 \right) \end{aligned} \quad (6)$$

Most equations for bedload transport empirically relate transport rate to a flow intensity, θ , (e.g., ASCE 1975, Chien and Wan 1999). As shown subsequently, the parameter $\theta d / k_i$ is convenient for indicating how cover roughness moderates θ . To simplify the discussion, the inflow and outflow rates of sediment, q_s , are taken as equal, thereby relaxing eq 6 to

$$\Pi_A = \varphi''_A \left(Re, Re_*, \eta \frac{d}{k_i}, S_0 \right) \quad (7)$$

which also recasts θ as $\eta = \theta / \theta_c$, thereby expressing θ relative to a critical value, θ_c , for incipient sediment movement. Many relationships for Π_A are expressed in terms of η or an excess of flow intensity, $\eta - 1$ (e.g., ASCE 1975).

The ensuing discussion considers how the parameters in eq 7 influence flow and sediment movement. It begins with a brief review of the pertinent cold-water properties.

Water-Temperature Effects

Ice is attended by cold water, usually at or slightly above 0°C. Most empirical relationships for alluvial-channel hydraulics are based on data obtained with water in the range of 10 to 20°C. All but one of the independent parameters in eq 7 directly involves water properties: v , ρ , and $\Delta\rho$. Reduced water temperature increases kinematic viscosity, v (it increases 100%, when water cools from 25 to 0°C), and slightly changes ρ (it increases about 0.3%, when water cools from 25 to 0°C, but attains a maximum at 4°C). An increase in v directly reduces Re and Re_* values, at constant q . In so doing, it increases flow drag on the bed, decreases particle fall velocity, and thereby overall increases flow capacity to convey suspended sediment. In many respects, the effect of low water temperature can be taken into account using Re , Re_* , and θ (insofar that it scales particle size and fall velocity relative to bed shear velocity, u_{*b}). The quantitative impacts of increased fluid viscosity on macro-turbulence are unclear as yet.

Quite a few studies have investigated water-temperature effects on sediment transport or, say, on sediment fall velocity. The studies confirm that sediment-transport rate increases with decreasing water temperature. Lane et al. (1949) and Colby and Scott (1965) show that trend in field data taken from the Missouri, Colorado, and Middle Loup rivers. Extensive flume experiments are reported by Ho (1939), Straub (1955), Colby and Scott (1965), Taylor and Vanoni (1972), and Hong et al. (1984). Taken together, the flume data confirm that sediment transport rates increase as water temperature decreases, the increases becoming substantial when water temperature drops below about 15°C. The flume data reported by Hong et al. (1984), for instance, show that the mean concentration of bed-sediment transport increased by factors up to seven and ten for a water temperature drop from 30 to 0°C. The increase, obtained with $d = 0.11$ mm, is attributable to increased concentration of sediment transport in a bed layer (layer thickness taken as $d\eta^{0.5}$) and increased uniformity of concentration distribution over the flow depth. The latter effect is largely attributable to the reduced fall velocity of suspended particles. Hong et al. (1984) concluded that temperature reduction significantly increases bed-level concentration of sediment movement only if bed-layer Reynolds number, Re_B (defined by Hong et al. as $[u_* d\eta^{0.5}] / v$) exceeds about 20; $Re_B = Re(\eta)^{0.5}$.

No study seems yet to have looked at the settling velocity of cohesive sediments, or cohesive-sediment behavior overall, at water temperatures close to 0°C. For example, Huang (1981), examined water-temperature effects on cohesive-sediment fall velocities for the range 32°C down only to 6.1°C.

Sediment Movement and Bedforms

The overall magnitude of the tractive force (drag and lift components) flow exerts on bed particles, together with the impacts of flow turbulence in all its scales, prescribe bed sediment motion. Ice-cover presence influences water drag on the bed and turbulence generation by redistributing flow and reducing the rate of flow energy expended along the bed. In so doing, cover presence poses three practical issues in using eq 6.

The first issue concerns estimation of τ_b or u_{*b} , shear stress or velocity associated with the channel bed. These variables are considerably more difficult to estimate than for open water loose-bed hydraulics.

A second issue is that the dependent loose-bed parameters (Π_A) of practical importance for alluvial-bed flows typically are estimated using semi-empirical relationships developed for open water conditions. Simply stated, at issue is the applicability of open water empirical relationships for ice-covered flow.

A third issue concerns the streamwise variation of the flow and sediment-transport capacity of an ice-covered channel. If the sediment-transport capacity of an ice-covered channel were less than the rate at which sediment load is supplied to the channel, the bed must locally aggrade. If the converse holds, the bed must degrade locally. The former condition usually would prevail for a free-floating cover, because the bulk velocity of flow decreases. The latter condition may occur when the cover is fixed or thick (large T/d), because the bulk velocity of flow under the cover increases. The various states of ice-cover condition complicate prediction of flow resistance and sediment transport in ice-covered alluvial channels.

A complicating aspect of estimating flow resistance and sediment transport is that the single relevant length scale for ice-covered flow is the total flow depth, Y , which itself usually is a dependent variable. For open water flow, flow drag on the bed can be characterized using Re and D_s , because they are not explicitly dependent on flow depth and flow velocity, depending instead on q as well as water and particle properties.

Two practical concerns are whether river ice influences bedform geometry and, if so, is its influence describable using relationships developed for open water flow. These issues have implications for estimating flow resistance and mixing processes. Following from eq 7, bedform length, L , and steepness, δ , can be expressed functionally as

$$L_* = L/d = \varphi_{L_*} \left(Re, Re_*, \eta \frac{d}{k_i}, S_0 \right) \quad (8)$$

and

$$\delta = H/L = \varphi_{\delta} \left(Re, Re_*, \eta \frac{d}{k_i}, S_0 \right) \quad (9)$$

in which H = bedform height. Equations 8 and 9 indicate that ice-cover presence should influence bedform geometry. The practical concern is accurate estimation of η , or u_{*b} . Figures 9 and 10 show that bedform geometry in ice-covered flow essentially conforms to the same relationships as prevail for open water flow. Figure 10 shows additionally that an ice cover, by reducing excess flow intensity at the bed, $\eta - 1$, reduces bedform steepness for the range of values indicated.

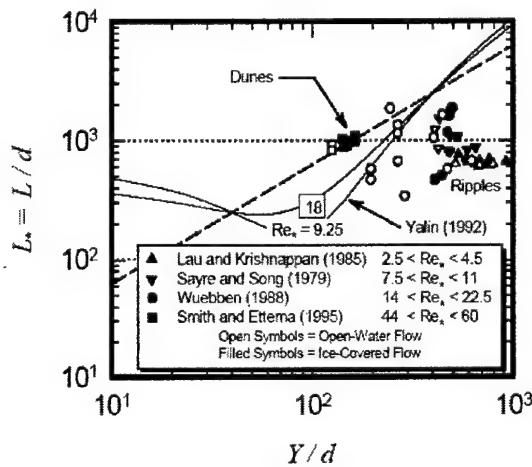


Figure 9. Flume data on bedform length in ice-covered flow conform to empirical open water curves developed by Yalin (1992).

However, there is an important cover influence not immediately evident from eq 8 and 9, and Figures 9 and 10. The influence is not adequately described in terms of cover influence on η or u_{*b} . Bedforms generate macro-scale turbulence, or coherent turbulence structures. Cover presence, by redistributing flow, influ-

ences the development of macro-turbulence and its consequences for bed sediment suspension as well as other dispersive processes. Recent experiments by Ettema et al. (1999) suggest that a smooth, level cover may invigorate macro-turbulence generation, mildly increasing the frequency of structures generated from bedforms and enabling them to penetrate the full depth of flow.

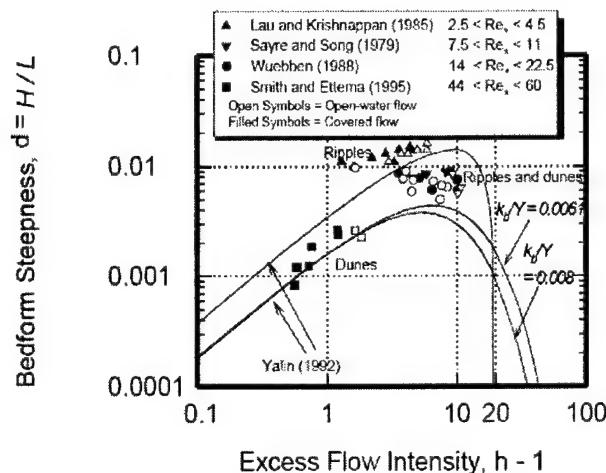


Figure 10. Flume data on bedform heights in ice-covered flow conform to empirical open water curves developed by Yalin (1992).

Flow Resistance in Alluvial Channels.

The issues for flow resistance hinge on the issues mentioned above for sediment entrainment, bedforms, and macro-turbulence. They entail estimation of resistance coefficients, f_b , associated with the bed and the ice cover, f_i , then estimation of flow depth, Y , given q . From eq 9

$$Y/d = Y_* = \varphi_Y \left(Re, Re_*, \eta \frac{d}{k_i}, S_0 \right). \quad (10)$$

Equation 10, however, is not immediately useful for predictive purposes, because open water methods estimate Y as the composite of form-drag and skin-friction resistance components. It is more helpful to use the Darcy-Weisbach re-

lationship, written in terms of unit discharge, q , of flow in a wide channel with a free-floating ice cover

$$Y_I = \left(\frac{f_i q^2}{4gS} \right)^{1/3} \quad (11)$$

with flow hydraulic radius $R_I = Y/2$

$$f_i = 0.5 f_b \left(1 + \frac{f_i}{f_b} \right),$$

and

$$f_b = f'_b + f''_b.$$

The functional relationship for each of these component resistance coefficients can be adjusted in terms of parameters used by existing empirical, estimation relationships.

For bed-surface resistance:

$$f'_b = \varphi_{f'_b} \left(Re, D_*, S_0, \frac{d}{k_i} \right) \quad (12)$$

For Form-drag resistance attributable to bedforms, such as dunes:

$$f''_b = \varphi_{f''_b} \left(Re, D_*, S_0, \frac{d}{k_i} \right) = \varphi_{f''_b} \left(Re, D_*, f'_b, \frac{d}{k_i} \right) \quad (13)$$

For the ratio:

$$f_i / f_b = \alpha = \varphi_\alpha \left(Re, D_*, S_0, \frac{d}{k_i} \right) = \varphi'_\alpha \left(Re, Re_{*b}, S_0, \eta \frac{d}{k_i} \right) \quad (14)$$

Again, an immediate practical issue implicit in eq 12 through 14 is that flow resistance in ice-covered alluvial channels can be estimated using open water re-

relationships, provided the influence of $\eta d/k_i$ in conjunction with the other parameters can be determined. If its influence can be determined, open water relationships, such as that given by Einstein and Barbarossa (1952) or Engelund and Hansen (1967), can be used to predict bed resistance in ice-covered loose-bed flow. A semi-empirical expression for eq 14 is given in Figure 11, which contains data from several flume studies.

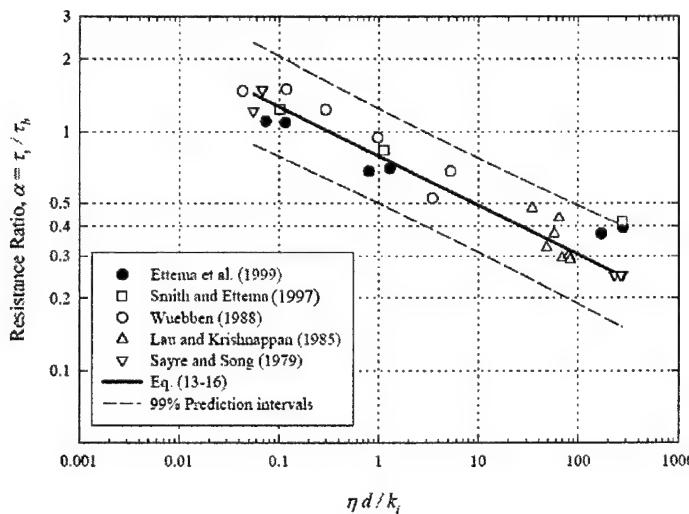


Figure 11. Resistance ratio, α , for an ice-covered flow in an alluvial channel.

Smith and Ettema (1997) developed a method, based on laboratory flume data, for estimating flow resistance in ice-covered alluvial channels. Their method is iterative and uses the following assumptions.

The mechanics of bedform formation essentially is the same for open water and ice-covered channels.

Methods for predicting bedform drag in open water flow (e.g., the Einstein-Barbarossa method, or the Engelund method) can be used to predict bedform drag in ice-covered flow. This can be done by replacing the bulk drag term, $\rho g Y_l S$, with an estimate of the actual bed shear stress in an ice-covered flow.

The ratio of boundary shear stresses along the bed versus along the cover underside is estimated as

$$\alpha = \tau_i / \tau_b = 0.84(\eta d / k_i)^{-0.20} \quad (15)$$

Equation 15 is an equation fitted to the flume data shown in Figure 11. The limits of the equation have yet to be determined for values of η beyond those indicated in Figure 11.

The proposed method requires the following input variables: cover roughness, k_i ; median bed-sediment diameter, d ; submerged specific gravity of bed sediment, $\Delta\rho/\rho$; unit discharge of water, q ; channel slope, S_0 ; and an initial guess at flow depth, Y_I (say, $Y_I \approx 1.2Y_0$). The procedure uses the Einstein-Barbarossa method for predicting bedform resistance and predicts values of flow depth, Y_I .

Bed-Sediment Transport

A basic issue concerns an imbalance between rate of bed-sediment supply to an ice-covered reach, q_s , and the sediment-transport capacity of that reach, q_{sl} . This issue involves the complex problem of spatially varied flow and sediment transport, with all its repercussions on local channel slope and morphology. The sediment-transport capacity of an ice-covered channel can be expressed functionally as

$$\frac{q_{sl}}{\sqrt{g(\Delta\rho/\rho)d^3}} = \varphi_{q_{sl}} \left(Re, Re_{*b}, S_0, \eta \frac{d}{k_i} \right) \quad (16)$$

This equation functionally characterizes bedload and suspended-load portions of bed-sediment transport. A fundamental issue relates directly to estimation of η or Re_{*b} . However, now cover influence on macro-turbulence becomes especially significant, because macro-turbulence affects sediment entrainment and suspension.

Laboratory Data on Bed-sediment Transport

When examined in terms of η , or u_{*b} , laboratory data on bedload capacity of ice-covered flow concur well with the open water trend shown in Figure 12 for Einstein's (1950) method and Meyer-Peter and Mueller's formulation (1948). Essentially, if η can be estimated, bedload transport in an ice-covered channel can be estimated using an open water method, such as the two used in Figure 12.

Estimation of suspended load in an ice-covered channel is not as straightforward as bedload estimation. Suspended load depends not only on the bed shear stress, or u_{*b} , but also on macro-turbulence and flow distribution. As mentioned above, cover presence likely significantly alters these. So far, there is no direct

way to account for macro-turbulence (notably the boils observed in sand-bed channels) effect on suspended load.

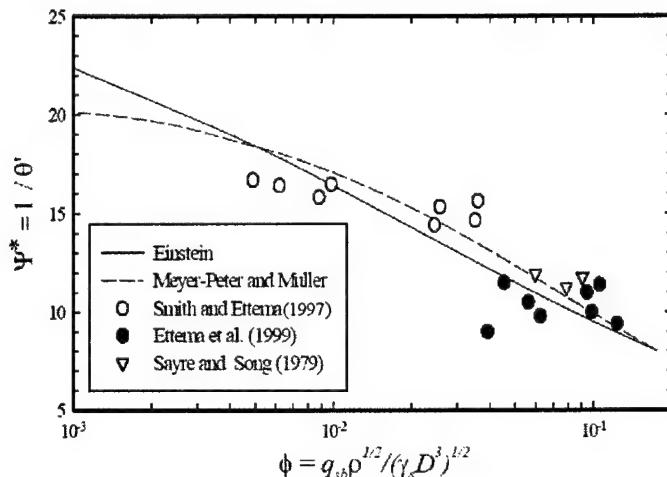


Figure 12. Bedload data compared with curves generated using Einstein's procedure and Meyer-Peter and Muller formula developed from open water data. Covered flow data conform to the same curves developed using open water data.

By virtue of its reduction of bulk velocity of flow, U , and thereby τ_b , and η , a free-floating ice cover typically reduces a channel's capacity to transport bed sediment. At certain zones within a channel, where the cover concentrates flow, sediment-transport rates may increase locally, however. Several laboratory studies have investigated cover-presence effects on sediment transport rate (Sayre and Song 1979, Wuebben 1986, Wuebben 1988a, Ettema et al. 2000). They all involved a free-floating cover that rises and subsides with changing flow rates. Their findings confirm that cover presence reduces rates of sediment transport. The rates decline rapidly with cover presence. Bedload-transport rate, for instance, can be almost halved by an ice cover that raises flow depth 15%, for a constant flow rate; this estimate assumes, reasonably, that bedload-transport rate $\propto \tau_b^2 \propto U^4$, with U decreasing by 15%; τ_b is shear stress acting on the bed, and U is bulk velocity of flow (ice-covered or open water). An important point here is that sediment eroded under an ice cover may not be transported far from the erosion location.

Field Data on Bed-sediment Transport

Few field studies have been conducted in which rates of sediment transport were measured for ice-covered channels. The studies indicate the inherent difficulty of obtaining such measurements and of interpreting them. Lawson et al. (1986) conducted an extensive study of flow and sediment movement at a reach of the Tanana River, Alaska. They obtained measurements of bedload and suspended-load rates at one cross section. The rates were comparable in magnitude to rates measured during a survey conducted about a year earlier at two cross sections close to that used by Lawson et al. Burrows and Harrold (1983) describe the earlier survey. Together, these data sets indicate a great reduction in the ratio suspended load relative to the bed-load from summer to winter. The reduction is attributed tentatively to reduced flow of melt water from glaciers drained by the Tanana River. Laboratory data obtained by Lau and Krishnappan (1985) and Ettema et al. (1999) show the opposite result, which both studies attribute to cover under-damping of turbulence generated by flow over bedforms.

Alterations in flow distribution complicate evaluation of ice-cover effects on transport rates for many. This difficulty is evident in Figure 1, which shows an ice cover over the Yellowstone River, near Fallon, Montana, and from figures such as Figure 4, which shows non-uniform ice accumulation across the Tanana River. The series of shear lines evident in the ice cover on the Yellowstone River (Fig. 1) indicate that the flow area has successively narrowed. Flow-width alteration is more difficult to predict than is flow depth change due to ice. The formation of subchannels within an ice-covered channel may accentuate narrowing of the flow area, especially if the channel is not prismatic. The subchannels form when accumulations of frazil slush or other ice pieces develop under the ice cover. In effect, they duct the flow in a manner that significantly alters the flow distribution from that attributable to the imposition of a level ice cover.

Field Data on Suspended Load

The few field studies on sediment transport during ice-covered flow focused on suspended-load and do distinguish between bed sediment and washload sediment.

The study carried out by Tywonik and Fowler (1973) focused on the measurement of suspended-sediment load in several rivers in the Canadian Prairie (e.g., Assiniboine River and Red River). They report that periods of ice cover on these rivers coincide with periods of low discharge and, therefore, low rates of suspended-sediment transport. In addition, they experienced considerable difficulty in making the suspended-load measurements, owing to frigid weather conditions and slush ice presence.

For most cold-regions rivers, the major sediment-transport event each year occurs during the large flows associated with ice runs resulting from the dynamic breakup of an ice-cover or the release of a breakup ice jam, if a jam develops. In addition to large flow rates usually involved, these events may produce severe gouging and abrasion of banks by moving ice. The resultant sediment transport comprises a mix of bed sediment and fine sediment washed into the river during snowmelt. Bedload measurements are very difficult to obtain under ice-run conditions.

Two studies, though, have provided some suspended-load data from individual breakup events. Prowse (1993) measured suspended-load concentrations during ice breakup of the Liard River, North West Territories. His data show a gradual increase in concentration with increasing water discharge immediately prior to breakup. When breakup occurred, suspended-load concentration increased by an order of magnitude, being comparable to concentrations associated with peak open water flows of about two to five times the peak flow at breakup. Data obtained by Beltaos and Burrell (2000) during ice breakup on the St John River, New Brunswick, show a similar trend.

Local Scour Beneath Ice Jams

The erosive behavior of a flow may increase locally beneath an ice jam if the jam concentrates flow, increasing the magnitude of its velocity and turbulence. Also, an ice jam may deflect flow, altering its direction in a manner so as to aggravate bank erosion or channel shifting. This mechanism locally increases flow velocity, and it may occur when flow and ice pieces are forced beneath an ice accumulation, such as an ice jam or an ice cover. Localized scour of an alluvial bed or bank of a channel may occur in the vicinity of an ice cover when the flow field at the cover locally increases flow velocities and, thereby, increases flow capacity to erode bed or bank sediment. There are several conditions under which this mechanism may operate.

The most severe condition typically occurs near the toe of an ice jam (freezeup or breakup), as illustrated in Figure 13. There, where jam thickness is greatest and flow most constricted, increased flow velocities may locally scour the bed (Neill 1976, Mercer and Cooper 1977, Wuebben 1988b). Channel locations recurrently (nominally every year) subject to ice jams may develop substantial scour holes. Tietze (1961) and Newbury (1982), for example, suggest instances of such scour holes at sites of recurrent freezeup jams. In most circumstances, the scour hole would have no lasting or adverse effect on channel morphology, because it gradually would fill once the jam is released. It is conceivable that, in certain circumstance nonetheless, the localized scour could have a

longer-term effect on channel morphology; e.g., if it promoted bank erosion at the jam site, or led to the washout of the channel feature triggering the jam, an island or bar, for example.

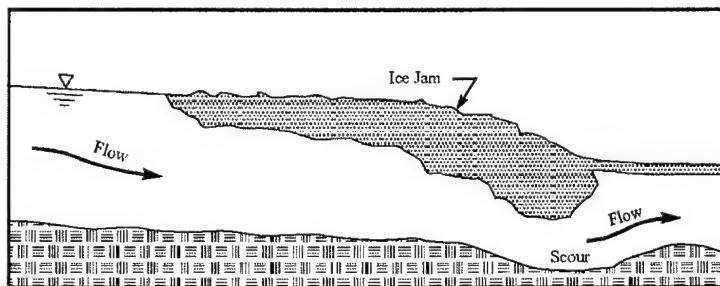


Figure 13. Flow acceleration and local scour beneath an ice jam.

To a lesser extent, local scour of bed and bank may occur also when ice pieces collect at the leading edge of an ice cover or at some channel feature (e.g., a set of channel bars) that impedes their drift. These situations are quite marginal in extent, likely occurring more-or-less randomly along a channel, and are short-lived. However, they potentially may trigger more severe erosion in some situations.

Towboat Activities

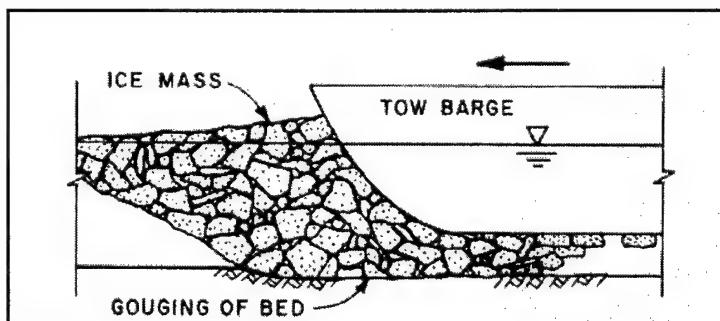


Figure 14. Channel bed gouged by ice rubble shoved by tow barge.

As illustrated in Figure 14, towboat activities in ice-covered water can mobilize bed sediment. The lack of clearance beneath a towboat or barge (draft of 2.7 m when fully laden) causes ice rubble to be shoved as an accumulated mass

ahead of it. Rubble ice is pushed down and may gouge the channel bed. If sufficient ice rubble accumulates, the towboat or barge even may become stuck on the mass of ice rubble.

5 ICE INFLUENCES ON CHANNEL MORPHOLOGY

The extents to which the seasonal formation and breakup of ice perturbs the stability of alluvial channels in regions subject to frigid winters are unclear. It seems from limited field observations (e.g., Mackay et al. 1974, Zabilansky et al. 2002) that river ice potentially may exert a compound impact of hydraulic and geomechanic effects that may continually destabilize certain plan-form geometries of channels subject to substantial, reservoir-regulated flow during frigid winters.

Introduction

There is considerable debate about ice influences on channel morphology and bank erosion. On one hand, some articles (e.g., Neill 1982, Blench 1986) largely dismiss the influences. On the other hand, there are fairly numerous anecdotal articles (e.g., Marusenko 1956, Lane 1957, Collinson 1971, MacKay et al. 1974, Hamelin 1979, USACOE 1983, Doyle 1988, Uunila 1997, Milburn and Prowse 1998), and the odd review article (Ettema 1999), suggesting ways in which river ice perceptibly affects channel morphology. The dismissive articles draw their conclusions from a few observations of overall plan form of a few rivers. They do not consider the impacts of reservoir regulation of flow during winter, take into account the diversity of channel morphologies, nor consider the important ephemeral impacts of ice that trigger local changes in thalweg.

Several factors influence alluvial channel stability. Most of them are explainable in terms of eq 16. Dependent variables of practical interest are average depth of flow, Y , hydraulic radius, R , also channel width, B , sinuosity, ζ , and shape, flow-energy gradient, S , and sediment-transport capacity, q_{sl} . Significant changes in any of the independent variables in eq 16 may alter R , ζ , or q_{sl} , and may destabilize the alluvial reach. The greatest natural disturbances typically result from changes in water and sediment inflow rates q , or q_s . River-ice impacts likely become more significant when water discharge fluctuates appreciably; then, the prospects for other adverse ice influences increase, such as ice-cover breakup followed by ice jamming.

Hydraulic Impacts

Ice may exert the following hydraulic impacts on alluvial channels.

By reducing the sediment-transport capacity of a river reach, ice may redistribute bed sediment along the channel. Whatever local effects river ice may ex-

ert, overall river ice usually reduces the channel's overall capacity to convey the eroded sediment a significant distance from the erosion location. Consequently, bars may develop in response to flow conditions under river ice, and be soon washed out shortly after the cover breaks up. In situations where a significant load of bed sediment enters a long reach that has a free-floating ice cover, river ice may tend to cause mild aggradation of the channel it covers. In situations where the reach is under a fixed ice cover, local degradation may occur.

Through its effects on lateral distribution of flow resistance and, thereby, flow and boundary drag, ice may modify channel cross-sectional shape developed under open water flow conditions.

Congestion or jamming of ice at one channel location may divert flow into an adjoining channel, which then enlarges (anabranching or thalweg avulsion), or over-bank, which may result in a channel cutoff (avulson).

Difficulties in ice passage through channel confluences may initiate ice jamming at confluences. In turn, an ice jam may modify confluence bathymetry.

By imposing additional flow resistance, a free-floating ice cover diminishes the effective gradient of flow energy available for sediment transport and alluvial-channel shaping. It consequently alters channel-thalweg alignment.

Ice jams, especially breakup ice jams, likely exert the greatest ice-hydraulic impact on unregulated alluvial channels. Mackay et al. (1974), for instance, describe the significant impacts that breakup ice jams exert on the Mackenzie River. For channels regulated by reservoirs used for hydropower generation during winter, ice-cover formation and presence can exert significant effects (e.g., Zabilansky et al. 2002). In overall terms, ice impacts have yet to be rigorously investigated, or even to be assessed quantitatively. Brief discussions of the impacts ensue.

Ice-Cover Influence on Local Elevation of Channel Bed

A basic issue concerns an imbalance between unit rate of sediment supply to an ice-covered reach, q_s , and the sediment-transport capacity of that reach, q_{sl} . This issue involves the complex problem of spatially varied flow and sediment transport, with all its repercussions on local channel slope and morphology. If the sediment-transport capacity of an ice-covered channel, q_{sl} , were less than the rate at which sediment load is supplied to the channel, q_s , the bed elevation must rise locally. Conversely, if $q_{sl} > q_s$, the bed elevation must drop locally. The former condition usually would prevail for a floating cover, because bulk velocity of flow decreases. The latter condition may occur beneath when the cover is fixed or thick, or both, because the bulk velocity of flow is forced to increase substan-

tially under the ice cover, with some flow spilling over the cover, as indicated in Figure 15. An ice jam, by constricting flow, may locally scour a riverbed, especially at the jam's toe (Neill 1976, Wuebben 1988b), as shown in Figure 13.

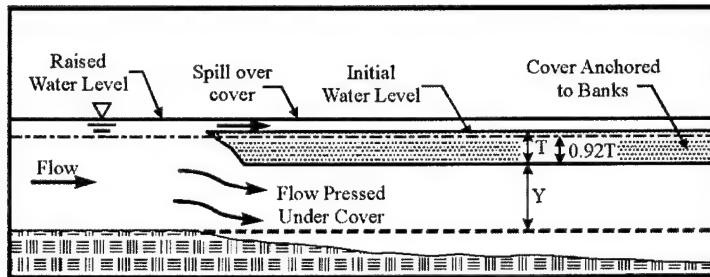


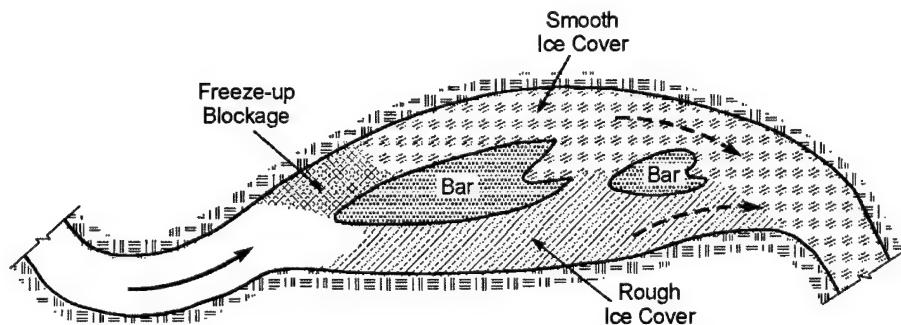
Figure 15. Flow in a channel reach constricted by a fixed ice cover.

Channel Anabranching, Avulsions, and Cutoffs

Channels with tight meander loops or with subchannels around numerous bars or islands are prone to ice-jam formation. Such channels typically have insufficient capacity to convey the incoming amount of ice. Their morphology may be too narrow, shallow, curved or irregular to enable drifting ice pieces to pass. Jam formation may greatly constrict flow, causing it to discharge along an alternate, less resistant course. Milburn and Prowse (1998), King and Martini (1984), and Dupre and Thompson (1979) suggest that ice-jam induced avulsion plays a major role in shifting the distributary channels of river deltas. Zabilansky et al. (2002) indicate that ice-induced avulsions of subchannels may occur in sinuously braided reaches of the Missouri River.

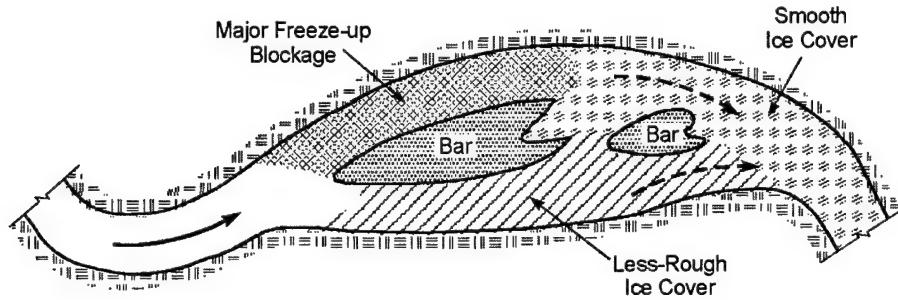
At sites where the river flows in two or more subchannels, ice-cover formation can trigger a switch of the principal thalweg from one subchannel to the other. Figure 16 illustrates the processes involved. When a rougher ice cover formed in one subchannel, the cover partially diverted flow from that subchannel to the one with the smoother ice cover. The subchannel with the smoother ice cover then enlarged while the rougher-covered one reduced in size. Survey observations from the Fort Peck reach of the Missouri River (Zabilansky et al. 2002) suggest that thalweg switching is a recurrent process, and that switches may take a period of several winters to fully occur. Strictly speaking, such switching composes a stochastic dynamic process that may be narrow-banded about a dominant period (e.g., a certain number of winters). It also may be broad-banded owing to the several factors (e.g., variability of flow conditions during a year or during ice-cover formation).

When an ice jam forms in a meander loop, upstream water levels may rise to the extent that flow proceeds over-bank and across the neck of a meander loop. If the meander neck comprises readily erodible sediment, and the flow is of sufficient scouring magnitude, flow diverted by the jam may result in a meander-loop neck cut, whereby a new channel forms through the neck, and the former channel is left largely cut off. A meander cut off shortens and steepens a channel reach, the consequences of which are felt upstream and downstream of the cut off reach. The net effect of ice jams, in this regard, is to reduce channel sinuosity. Mackay et al. (1974), for instance, cite examples of such events.



a. A relatively short initial accumulation of drifting ice in the top subchannel (1) may divert ice into the bottom subchannel (2), which then becomes extensively enveloped by a rough ice cover. Meanwhile, subchannel 1 freezes over with a smooth ice cover, or may remain partially open. The greater flow resistance in subchannel 2 causes flow to favor subchannel 1, which then enlarges.

Figure 16. Ice-cover formation in a sinuous-braided channel may alternate the location of the major subchannel. Two scenarios for alternation of major subchannel were identified.



b. A relatively long initial accumulation of drifting ice in the top subchannel (1) may divert ice and flow into the bottom subchannel (2), which then becomes extensively enveloped by a less rough ice cover. The greater flow resistance in subchannel 1 causes flow to favor subchannel 2, which then enlarges.

Figure 16 (cont'd).

If, on the other hand, the meander loop is wide and not easily eroded, over-bank flow resulting from an ice jam may have the reverse effect. Rather than the net consequence being the erosion of channel through the meander loop, over-bank flow may deposit sediment, thus raising bank height and reinforcing the meander loop. Eardley (1938) reports that ice jams cause substantial sediment deposition on the flood plain of the Yukon River. A similar event is reported in Simon et al. (1999) for the Fort Peck reach of the Missouri River. Over-bank deposition of sediment, together with ice-run gouging and abrasion of sediment erosion from the lower portion of a bank, may over-steepen riverbanks.

Channel Confluences

By virtue of their role in connecting channels and, thereby, concentrating ice within a watershed, confluences are perceived as locations especially prone to the occurrence of ice jams. Fairly numerous accounts exist of jams in the vicinity of a confluence (Tuthill and Mamone 1997). Flow and ice concentration in a confluence may cause ice to jam within a confluent channel, within the confluence itself, or at some distance downstream of a confluence. Various mechanisms may trigger jams in the vicinity of confluences. Confluence bathymetry commonly leads to jam initiation, and in turn jamming can modify confluence bathymetry (Ettema and Muste 2001).

Cover Influence on Thalweg Alignment

An ice cover reduces the effective energy gradient of flow (and thereby stream power) available for sediment transport and channel shaping. Therefore, cover formation may trigger a change in thalweg alignment. Such an influence still is a matter of conjecture, and indeed is difficult to confirm from field or laboratory experiments, because of the long durations involved. Zabilansky et al. (2002) present a hypothesis that elaborates this influence.

The essential consideration here is that thalweg lengthening and branching are mechanisms whereby an alluvial-channel flow innately increases flow resistance (and thereby rate of energy use) to offset increased flow energy associated with a larger channel slope.

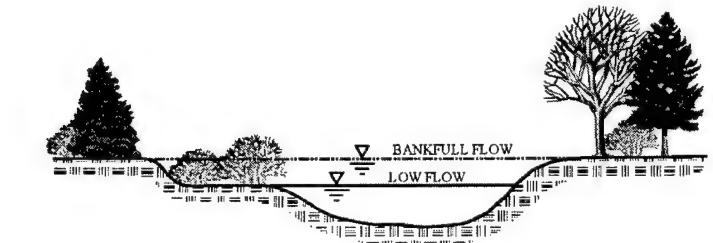
Impacts on Riverbanks

Ice may influence channel cross-section shape, alignment, and bed elevation through several geomechanic impacts on riverbanks:

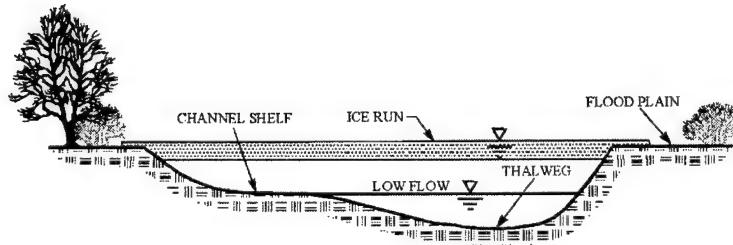
- Reduce riverbank strength by increasing pore-water pressure or by producing rapid drawdown of bank water table during dynamic ice-cover or ice-jam breakup. This impact is part of the overall consequence of freeze-thaw behavior or riverbanks in frigid conditions.
- Tear, batter, and dislodge riverbank material and vegetation during collapse of bank-fast ice.
- Gouge and abrade riverbank material and vegetation during ice run.

These impacts reduce riverbank resistance to erosion and increase the local supply of sediment to the channel. The first two impacts are not well studied. The third has received some attention, but the extent to which it affects channel shape is unclear. It is normal for river channels and floodplains subject to ice to be denuded of larger vegetation, as is sketched in Figure 17.

Engelhardt and Waren (1991), for instance, briefly describe the consequences of such combined processes for the Missouri River in downstream of dams in Montana and North Dakota—increased rates of ice-covered flow, increased movement up and down riverbanks, bank freezing at higher elevation, and more frequent freeze-thaw cycles exacerbating bank erosion. The consequences become noticeable in early spring, when noticeably large portions of riverbanks fail. Zabilansky et al. (2002) report similar observations.



a. Rivers not subject to ice.



b. Rivers subject to ice.

Figure 17. Ice runs may inhibit vegetation growth along riverbanks and floodplains.

Freeze-Thaw Influences on Riverbank Strength

It is well known that the freezing and thawing of soil affect the erosion of riverbanks adjoining rivers and lakes. Lawson (1983, 1985) and Gatto (1988, 1995), among others, provide extensive reviews of the subject. In short, because frozen soil is more resistant to erosion than is unfrozen soil, riverbanks are less erodible while frozen. The freezing and thawing of soil, however, usually weakens soils, making thawed (or thawing) riverbanks more susceptible to erosion. The net consequence on the overall rate of riverbank erosion, therefore, remains a matter of debate. Most likely, the net consequence varies regionally and from site to site.

Freeze-thaw cycles affect soil structure, porosity, permeability, and density. These changes in soil properties can substantially reduce soil shear strength and bearing capacity; strength reductions of as much as 95% are reported (Anderson and Anderson 1990). Such adverse effects on soil strength depend on soil-particle size and gradation, moisture content, the number and duration of freeze-

thaw cycles, and several other factors. Though there is no single, standard test to determine whether a soil is prone to significant weakening due to freeze-thaw (Chamberlain 1981), particle size is commonly used as an approximate indicator of soil sensitivity to freeze-thaw weakening. Soils containing fine sands and silts are especially sensitive, because they are permeable and susceptible to change in soil structure. By virtue of their particle size (about 0.1 to 0.06 mm [0.004 to 0.002 in.]), and the surface-tension property of water, fine sandy and silty soils absorb moisture more readily than do coarser or fine sediment. Clay soils are less sensitive, because of their low permeability. The variability of soil properties along a riverbank, and within a specific riverbank location, causes the effects of riverbank freezing to differ along a reach.

Gatto (1995) suggests that an eroding riverbank is especially subject to deep penetration of freezing, thereby making more of the riverbank prone to freeze-thaw weakening and erosion. The absence or stunted extent of vegetation that characterizes many eroding riverbanks results in diminished insulation protection of the riverbank and increased heat loss to air. In addition, the crest region of a riverbank experiences the greatest heat loss, owing to the crest's exposure to air on at least two sides. Because of its exposure to wind, the crest may also accumulate less snow. Less snow, in turn, means deeper frost penetration during winter and faster thaw in spring. However, less snowmelt would be available to percolate into the riverbank. Questions exist about the exact manner in which border ice is anchored to the riverbank, and other factors (notably, variations in water-table [or piezometric] surface and moisture content of the top zone of the riverbank) would modify the extent of the frozen zone and its connection with river ice. Presumably, if the top portion of the riverbank and upland were dry, the riverbank crest might be the zone of least heat loss, as the distance between air and water table is greatest there.

As the upper zone of frozen ground thaws, melt water likely drains down, over the surface of the still frozen ground. The riverbank, weakened by thaw expansion of ground and subject to the seepage pressures, is at its least stable, annual condition.

Several studies (e.g., Harlan and Nixon 1978, Reid 1985) have found that south-facing riverbanks experience lesser thickness of freezing, all else being equal, than do north-facing riverbanks. The explanation is that south-facing riverbanks (in the northern hemisphere) receive more insolation (energy in the form of short-wave radiation from the sun). South-facing riverbanks also may undergo more frequent diurnal freeze-thaw cycles (Gatto 1995, Zabilansky et al. 2002). The net effect on weakening of riverbank material of riverbank alignment has yet to be determined.

Reduction of Riverbank Strength

Flow stage and stage fluctuations influence seepage pressures and the freeze-thaw behavior of riverbanks. Higher flow stage raises the water table in a riverbank, and a rapid drop in flow stage may momentarily reduce riverbank stability by increasing seepage pressures and, thereby, reducing the shearing resistance of the material composing the riverbank. Ice-cover formation raises flow stage, whereas cover breakup may abruptly lower it. River-ice formation, thereby, may weaken riverbanks.

Riverbank freezing is closely linked to bankfast-ice formation along a channel, though the details of the relationship between them are unclear. They depend on riverbank condition (material, vegetation, snow, etc.), the relative elevations of water table and flow stage, and temperatures of groundwater and river water. The strength of bankfast-ice attachment to a bank depends on the relative elevations of the water table and flow stage, and on the relative water temperatures. A relatively warm (i.e., several degrees above the freezing temperature) flow of groundwater into the river will retard bankfast-ice growth and weaken its hold on the bank. The growth of a thick fringe of bankfast ice, on the other hand, may affect seepage flow through the bank, possibly constricting it, and slightly raising the water table. They are especially significant for regulated rivers, for which flows do not diminish during winter.

Bankfast-ice Loading of Bank

Bankfast-ice weakening of banks likely is significant for steep banks, typically those banks comprising sufficient clay as to be termed cohesive. It also likely is significant for banks whose water table declines in elevation away from flow elevation in a channel, because the bankfast ice is less securely anchored into the bank. This erosion mechanism seems not to have been investigated heretofore but was observed—e.g., along the Fort Peck reach of the Missouri River (Zabilansky et al. 2002). When the flow stage in a channel drops, portions of an ice cover attached to a bank during the higher flow stage may be left momentarily cantilevered from the bank. The cantilevered ice soon collapses, weakening and wrenching bank material as it does so.

Figure 18 illustrates how bankfast ice might weaken a bank. The ice cover freezes into the bank. The extent of the root is limited by groundwater elevation and temperature, and by the nature of the bank material. When the water level in the channel drops, and the ice cover breaks up, ice attached to the bank is cantilevered out from the bank, rotates, and tears a portion of the bank as it drops. It is difficult to get direct field observations of this mechanism for bankfast ice attached to vertical banks. For the moment, evidence for it is circumstantial.

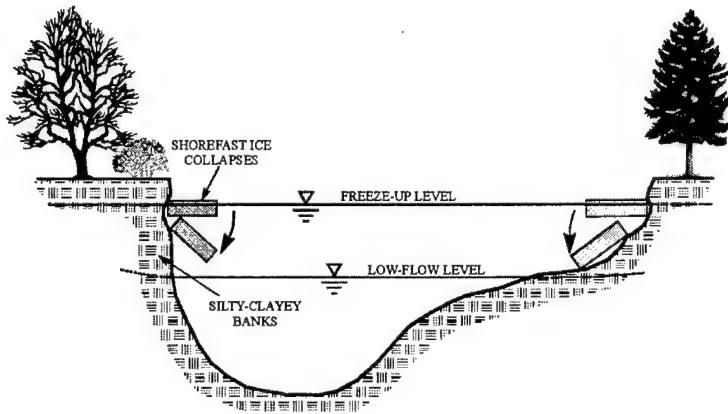


Figure 18. Collapse of shorefast ice may erode banks when flow stage is lowered.

Gouging and Abrasion of Banks

During heavy ice runs resulting from ice-cover breakup or ice-jam release, large pieces of ice potentially may gouge and abrade channel banks. There exists significant evidence showing that it substantially affects channel-bank morphology subject to dynamic ice runs (Marusenko 1956, Walker 1969, Smith 1979, Hamelin 1979, Rosen 1979, Martinson 1980, Uunila 1997, Brooks 1993, Wuebben 1995, Wuebben and Gagnon 1995). Such channels usually are relatively steep and convey high-velocity flows. Moreover, their ice covers typically break up fairly dramatically in concert with a sudden rise in flow, attributable, for example, to rapid snowmelt or rain, or both. The resultant ice rubble is composed of hard, angular blocks of ice.

One study of 24 rivers in Alberta (Smith 1979) led to the intriguing hypothesis that ice runs enlarge channel cross sections at bank-full stage by as much as 2.6 to 3 times those of comparable flow rivers not subject to ice runs. The hypothesis is based on a comparison of the recurrence interval of bank-full flows in the 24 rivers and an empirical relationship between the cross-section area and flow rate for bank-full flow. The channel-widening effect of ice runs is plausible. However, the extent of widening indicated seems overly large, and requires further confirmation. Kellerhals and Church (1980), in a discussion of Smith (1979), argue against Smith's hypothesis. They suggest that other factors have led to an apparent widening of the channels analyzed by Smith—e.g., recent entrenchment of major rivers in Alberta, and ice-jam effects of flow levels. Moreover, it is possible that the banks are somewhat protected by a band of ice forming a shear wall flanking the riverbanks. It is interesting to contrast Smith's hypothesis with a

further hypothesis mentioned above that ice jams may promote channel narrowing by causing over-bank flow (e.g., Uunila 1997). For channels whose dominant channel-forming flow coincides with ice-cover breakup, over-bank loss of flow reduces the flow rate to be accommodated by the channel.

In many situations, notably those in which an ice run is sluggish, a shear wall of broken ice may fend moving ice from contacting the bank (Fig. 19). The shear wall usually becomes smooth-faced, and protects riverbanks from direct ice impact or gouging. Running ice, if sufficiently thick, may still gouge the lower portion of a bank. Significant gouging may occur downstream of the toe of a jam, before the arrival of sufficient ice rubble to form shear walls. A surge front from a released jam may fracture an ice cover into large slabs, which then are set in motion. The surge front typically moves faster than the ice rubble composing the jam, but gradually attenuates. Typically, ice gouging occurs within a relatively short reach of a river.



Figure 19. Ice shear wall along riverbank.

Ice gouging and abrasion, though, can be severe for channel features protruding into the flow. In addition, channel locations with a substantial change in

channel alignment are especially prone to ice-run gouging and abrasion; e.g., a sharp bend, point bar, and portions of a channel confluence. There is a little information on how ice runs affect the local morphology of these sites. Two features have been observed in gravelly rivers—ice-push ridges and cobble pavements. Ice-push ridges form when a heavy ice run gouges and shoves sediment along the base of banks (e.g., Bird 1974). The gouged sediment piles up as ridges beneath the ice run as it comes to rest as a jam. The finer sediments eventually get washed out, leaving the more resistant gravel and boulders in ridges. The ridges usually develop in the vicinity of locations subject to recurrent ice jams.

Cobble pavements may cover bars and the lower portion of banks subject to ice gouging and abrasion. Essentially, an over-riding mix of ice and cobbles removes the finer material from the surface of the bars or banks. The resultant cobble surface comprises cobbles whose major axis is aligned parallel to the channel and whose size gradually decreases downstream (Mackay and Mackay 1977). The resultant cobble pavement may extend for many miles along the banks of large northern rivers, such as the Mackenzie and Yukon Rivers (Kindle 1918, Wentworth 1932).

The gouging and abrasion of the lower portion of banks, in conjunction with over-bank sediment deposition during ice-jam flooding, may produce an elevated ridge or bench feature along some northern rivers. These features have been dubbed bechevniks for Siberian rivers (Hamelin 1979). A bechevnik is the marginal strip composing the lower portion of a riverbank and exposed portion of adjoining river bed that, in days gone by, formed convenient paths for towing boats upstream manually or by horse; becheva apparently is Russian for towrope. Figure 20 illustrates the main features of a bechevnik, which may form partly from ice abrasion and partly from the deposition of sediment and debris left by the melting of ice rubble stranded after ice runs.

Ice-run gouging and abrasion have an important, though as yet not quantified, effect on riparian vegetation that, in turn, may affect bank erosion and channel shifting. Where ice runs occur with about annual frequency, riparian vegetation communities have difficulty getting established. Ice abrasion and ice-jam flooding may suppress certain vegetation types along banks, as illustrated in Figure 17 and in Figure 20 for a bechevnik, possibly exacerbating bank susceptibility to erosion. This aspect of river ice has yet to be further investigated. Scrimgeour et al. (1994) and Prowse (2001) provide useful early reviews.

Moving ice also may grind banks formed of soft rock (e.g., sandstones and mudstones) or stiff clay. Danilov (1972) and Dionne (1974), e. g., describe how moving ice has affected rock banks of rivers like the St Lawrence River. The extent of erosion, though, is less than for banks formed of alluvial sediment.

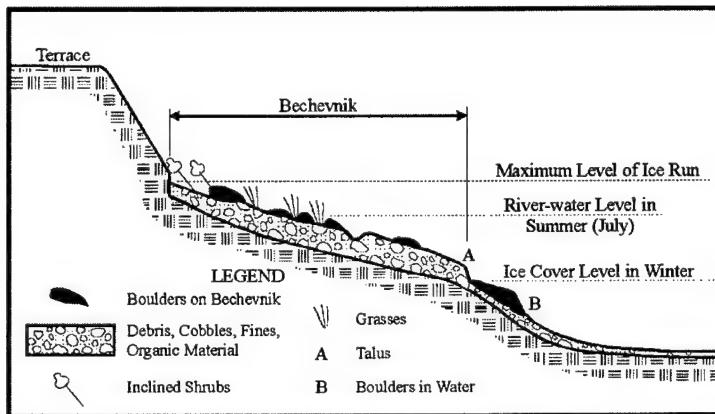


Figure 20. Bechevnik (after Hamelin 1979).

Combined Impacts

One hydraulic or geomechanic impact of river ice may disturb a channel, but not necessarily destabilize it. A combination of hydraulic and geomechanic impacts, though, may destabilize a channel. A shift in thalweg alignment or a bank failure, alone, may not destabilize a channel. The channel may adjust back to more or less its stable open water condition once open water conditions resume.

Channels usually considered less stable in open water conditions are more likely to be adversely affected by river ice. Sinuous-point-bar, sinuous-braided, and braided alluvial channels are especially prone to river ice impact, especially if they have steep banks, formed of fine and partially cohesive sediments. The thalwegs of such channels usually lie close to the outer banks of bends, and the banks themselves are prone to bankfast-ice loading, lack of vegetation cover (typical of eroding banks), and freeze-thaw weakening. Figure 21 illustrates this susceptibility. The thalweg lies close to the bank, such that the flow continually erodes the bank-toe, thereby keeping the bank steep, and possibly undercutting the bank. Snow cannot protectively blanket the bank face. Frost penetration potentially is deep, the water table is held relatively high, and the channel shifts, destabilized.

An intriguing question is whether the destabilizing impacts of river ice uniquely modify alluvial-channel morphology. An answer can only be tentatively suggested at this moment. It is likely that the major geometric parameters do not change appreciably (e.g., channel thalweg sinuosity, width, hydraulic radius, meander radius). However, river ice likely increases irregularities in channel plan-

form and the frequencies with which channel cross section and thalweg alignment shift.

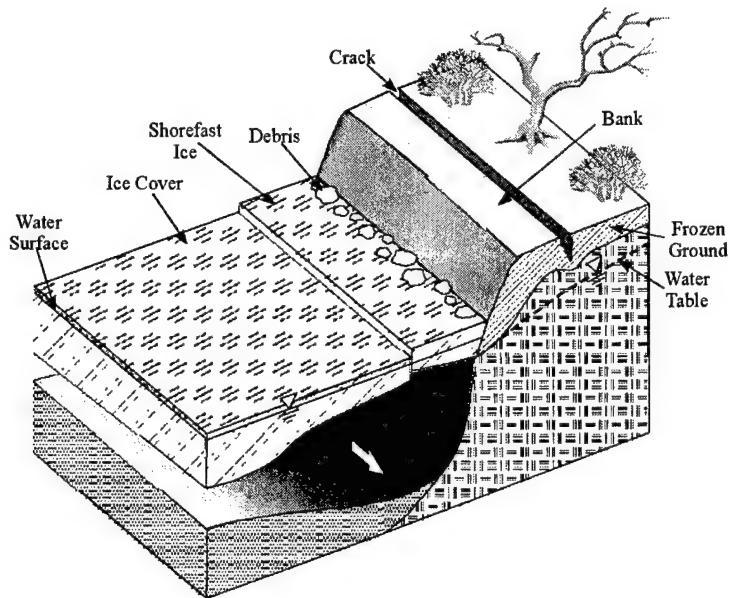


Figure 21. Combined hydraulic impacts (e.g., thalweg shift and bank-toe erosion) and geomechanic impacts (e.g., freeze-thaw weakening of bank material, elevated seepage pressures, bankfast-ice loading) may weaken and erode channel banks, especially along channel bends, and result in continual, overall channel destabilization.

6 SUMMARY

The annual cycle of ice formation and breakup influences sediment-transport dynamics in waterways. The magnitude of the influences depends on a combination of factors. Of prime importance are factors determining the amount of ice formed and the quantities of water and sediment to be conveyed.

The overall amount of ice is usually governed by the cumulative period of temperature degrees below the freezing temperature of water. Water-flow rate directly determines quantities of sediment transport. Under natural conditions in many waterways, the annual cycle of ice formation is accompanied by a decline in water runoff and channel flow. Rates of sediment supply and channel transport diminish commensurately. Runoff and channel flow subsequently increase during spring thaws, and it is then that ice-cover effects on sediment transport can become significant. For flow-regulated rivers and channels downstream of reservoirs, though, ice effects on sediment transport and alluvial-channel behavior are of special interest. Substantial flows may occur while such rivers and channels are ice covered in winter.

This report considered the following topics related to sediment transport:

- Ice-cover influences on flow distribution.
- Sediment transport by ice (i.e., sediment included in drifting ice).
- Sediment transport under ice.
- Ice influences on channel morphology.

Ice-Cover Influence On Flow Distribution

An ice cover imposes an additional resistant boundary that decreases a channel's flow capacity and vertically redistributes streamwise velocity of flow in a channel. If the cover is free-floating, it may reduce the erosive force of flow in the channel and thereby reduce rates of sediment transport. However, cover presence also may laterally redistribute flow, usually concentrating it along a thalweg. If the thalweg lies close to one side of a channel, flow concentration may locally increase bank erosion and channel shifting. On the other hand, if the thalweg is more-or-less centered in a channel, the cover may reduce bank erosion and channel shifting. Additionally, if the full cover were fixed to the riverbank, it may increase locally flow velocities and rates of sediment transport. The variability of flow response to ice-cover presence makes it difficult to draw simple overall conclusions about ice-cover effects on a river's bed and banks. The net effects will vary from site to site.

Sediment Transport by Ice

Sediment-laden ice slush and clumps of ice-bonded sediment may appear during the early stages of ice formation in certain rivers and streams subject to the winter cycle of ice formation. The ice slush and clumps usually include a mix of frazil and anchor ice that were once briefly bonded to the beds of such rivers and streams. The amounts of sediment entrained and rafted with the ice slush and clumps can produce a substantial momentary surge in the overall quantity of sediment moved by some rivers and streams. Much of the entrained sediment becomes included in an ice cover, where it remains stored until the cover breaks up. Though ice rafting of sediment is known to happen, the engineering and environmental implications of its occurrence are largely unknown. The mechanisms whereby ice entrains and transports sediment are not well understood. Also, the distances over which ice-rafted sediment typically may be transported are not really known.

Sediment Transport Under Ice

Sediment-transport processes in ice-covered flow are also not fully understood. Some influences of an ice cover on flow distribution are reasonably well understood, while some are barely recognized (e.g., on large-scale turbulence); few reliable methods exist for estimating transport rates or ice effects on channel stability. An essential feature of alluvial rivers and channels is that their morphology and flow-resistance behavior vary interactively with flow and sediment conditions. Depending on flow magnitude, ice covers modify the interaction, doing so over a range of scales in space and time.

Water-Temperature Effects

Ice is attended by cold water, usually at or slightly above 0°C. Most empirical relationships for alluvial-channel hydraulics are based on data obtained with water in the range of 10 to 20°C. Reduced water temperature increases kinematic viscosity, and slightly changes density. In so doing, it increases flow drag on the bed, decreases particle fall velocity, and thereby overall increases flow capacity to convey suspended sediment.

Sediment Movement and Bedforms

The overall magnitude of the tractive force (drag and lift components) flow exerts on bed particles, together with the impacts of flow turbulence in all its scales, prescribe bed sediment motion. Ice-cover presence influences water drag

on the bed and turbulence generation by redistributing flow and reducing the rate of flow energy expended along the bed.

Local Scour Beneath Ice Jams

The erosive behavior of a flow may increase locally beneath an ice jam if the jam concentrates flow, increasing the magnitude of its velocity and turbulence. Also, an ice jam may deflect flow, altering its direction in a way that aggravates bank erosion or channel shifting. This mechanism locally increases flow velocity, and it may take place when flow and ice pieces are forced beneath an ice accumulation, such as an ice jam or an ice cover. Localized scour of an alluvial bed or bank of a channel may occur in the vicinity of an ice cover when the flow field at the cover locally increases flow velocities and, thereby, increases flow capacity to erode bed or bank sediment.

Ice Influences on Channel Morphology

The extents to which the seasonal formation and breakup of ice perturbs the stability of alluvial channels in regions subject to frigid winters are unclear. It seems from limited field observations that river ice potentially may exert a compound impact of hydraulic and geomechanical effects that may continually destabilize certain plan-form geometries of channels subject to substantial, reservoir-regulated flow during frigid winters.

Impacts on Riverbanks

Ice may influence channel cross-section shape, alignment, and bed elevation through the following:

- Reduce riverbank strength by increasing pore-water pressure or by producing rapid drawdown of bank water table during dynamic ice-cover or ice-jam breakup. This impact is part of the overall consequence of freeze-thaw behavior of riverbanks in frigid conditions.
- Tear, batter, and dislodge riverbank material and vegetation during collapse of bank-fast ice.
- Gouge and abrade riverbank material and vegetation during ice run.

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One hydraulic or geomechanical impact of river ice may disturb a channel, but not necessarily destabilize it. A combination of hydraulic and geomechanical impacts, though, may destabilize a channel. A shift in thalweg alignment or a bank failure, alone, may not destabilize a channel. The channel may adjust back to more or less its stable open water condition once open water conditions resume.

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An intriguing question is whether the destabilizing impacts of river ice uniquely modify alluvial-channel morphology. An answer can only be tentatively suggested at this moment. It is likely that the major geometric parameters do not change appreciably (e.g., channel thalweg sinuosity, width, hydraulic radius, meander radius). However, river ice likely increases irregularities in channel planform and the frequencies with which channel cross section and thalweg alignment shift.

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